STRATIGRAPHY AND PALEOGEOGRAPHY OF LATE CRETACEOUS AND PALEOGENE ROCKS OF SOUTHWEST UTAH

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ABSTRACT

The Late Cretaceous to Paleogene sedimentary rocks of southwest Utah record active Sevier-style deformation, the cessation of Sevier tectonism, and the evolution of Laramide-style deformation. Three temporally overlapping tectonic episodes are recorded in these rocks and include: 1) active Sevier-style, thin-skinned thrust activity and foreland sedimentation; 2) postorogenic uplift and sedimentation of the thrust belt, and; 3) active Laramide-style basement-cored uplifts and folding.

The early Campanian upper Iron Springs and middle-to-late Campanian Kaiparowits formations represent synorogenic, fluvial sedimentation derived from the Sevier fold and thrust belt. The Iron Springs Formation received sediment from Precambrian to Upper Paleozoic strata exposed in the Wah Wah and Blue Mountain thrust sheets of southwestern Utah. Volcanic, radiolarian argillitic, and rare metamorphic lithic grains in the Kaiparowits Formation suggest a source in southeastern California and southern Nevada.

The late Campanian to early Paleocene Canaan Peak Formation was deposited in an east to northeast directed, braided fluvial system. Petrographic and geochemical analysis of volcanic and siliciclastic clasts indicate that the Canaan Peak and Kaiparowits formations were both derived from two different source terranes. The volcanic component was probably derived from the Jurassic Delfonte Volcanics of southeastern California. Siliciclastic detritus was derived from the Mississippian Eleana

Formation of southern Nevada. Mixing of these two provenances occurred between the Spring Mountains of southern Nevada and the Pine Valley Mountains of southwestern Utah.

The early Paleocene Formation of Grand Castle (informal name) has previously been mapped as the basal Claron Formation. The Formation of Grand Castle stratigraphically lies between the Canaan Peak and Pine Hollow formations in the Table Cliff Plateau, suggesting a formational status for the Grand Castle. The Formation of Grand Castle represents an east to southeast flowing braided river system. Clast and sandstone lithologies suggest that the Formation of Grand Castle had the same provenance as the Iron Springs Formation (the Wah Wah and Blue Mountain thrust sheets).

The laterally extensive conglomerates of the Canaan Peak Formation and Formation of Grand Castle may represent a northward progression of postorogenic isostatic uplift and erosion of the Sevier fold and thrust belt. The Formation of Grand Castle truncates the easternmost Sevier thrust faults, substantiating a post-Sevier origin for the Grand Castle conglomerates.

The early Paleocene to middle Eocene Pine Hollow Formation unconformably overlies both postorogenic sequences and records active Laramide partitioning of the foreland basin. The Pine Hollow basin received sediment from both the west and northeast, and is associated with the development of the Johns Valley anticline and possibly the Circle Cliffs uplift. Small alluvial fans developed on the limbs of these structures and grade laterally into sheet-flood sandstones, playa-mudflats, and lacustrine limestones in the center of the basin.

Fluvial, deltaic, and lacustrine deposits of the Claron Formation overlap paleotopographic highs of the Pine Hollow basin. This overlap assemblage indicates cessation of Laramide deformation by the middle Eocene. The Claron Formation is a time-transgressive sequence with a basal age of late Paleocene in the west (Pine Valley Mountains) and a middle Eocene age to the east (Table Cliff Plateau). The upper age limit is early Oligocene. Lacustrine facies transgressed to the north and northeast over relatively flat, pedogenically altered floodplain deposits.

STRATIGRAPHY AND PALEOGEOGRAPHY OF LATE CRETACEOUS AND PALEOGENE ROCKS OF SOUTHWEST UTAH

INTRODUCTION

During the Late Cretaceous and early Tertiary, two episodes of contractional deformation in the western Cordillera controlled the patterns of sedimentation within Utah (Armstrong and Oriel, 1965; Armstrong, 1968; Fouch et al., 1983; Lawton, 1983; Heller et al., 1986; Decelles, 1988). During the Late Cretaceous, thin-skinned deformation (the Sevier orogeny) formed highlands of moderate relief from which sediment was shed eastward into the foreland basin (Spieker, 1946; 1949; Armstrong, 1968; Fouch et al., 1983; Lawton, 1983). During the latest Cretaceous and Early Tertiary, basement-cored uplifts (the Laramide orogeny) partitioned the Sevier foreland basin, creating internally drained basins dominated by low-energy fluvial and lacustrine deposition (Chapin and Cather, 1981; Lawton, 1983; Cross, 1986; Dickinson et al., 1986; Dickinson et al., 1988; Lawton and Trexler, 1989).

Late Cretaceous and Paleogene sedimentary rocks of southwestern Utah represent nonmarine sedimentation which spans the transition from Sevier to Laramide style deformation. Two laterally different and petrographically distinct coarsening-upward sequences record cessation of Sevier-style tectonism (Goldstrand, 1990a). These two sequences are; 1) the middle

Campanian to early Paleocene Kaiparowits-Canaan Peak formations and, 2) the Late Cretaceous to early Paleocene Iron Springs-Grand Castle (informal name) formations.

Heller et al. (1988) present a model suggesting that fine-grained intervals in a coarsening-upward foreland sequence represent active thrusting, whereas coarse-grained deposits represent cessation of thrust activity. According to Beaumont (1981) and Jordan (1981) foreland basin subsidence is most rapid during active thrust emplacement. Coarse, synorogenic sedimemts are deposited adjacent to the thrust front, whereas finer grained sediments are deposited in the distal foreland (Heller et al., 1988). Cessation of thrust activity and removal of the thrust sheets by erosion, results in flexural rebound of the thrust belt and proximal foreland basin (Heller et al., 1988). During this postorogenic phase of uplift, a sheet of coarse clastic detritus is deposited in the distal foreland basin and overlies finer grained synorogenic deposits (Heller et al., 1988; Paola, 1988).

The early Paleocene to middle Eocene Pine Hollow Formation unconformably overlies the post-Sevier Canaan Peak Formation and the Formation of Grand Castle, and represents Laramide partitioning of the foreland basin (Goldstrand, 1990a). Fluvial, deltaic, and lacustrine deposits of the Claron Formation gradationally overlies the Pine Hollow Formation and overlap and post-date Laramide structures (Goldstrand, 1990a). The Claron Formation is a time-transgressive sequence ranging in age from late Paleocene to early Oligocene.

This paper describes the stratigraphy of the Late Cretaceous and Paleogene nonmarine strata of southwestern Utah. Paleocurrent analysis, depositional system reconstructions, and provenance studies are used to reconstruct the paleogeography from early Campanian to middle Eocene. These data document three temporally overlapping tectonic episodes recorded in the Late Cretaceous and Paleogene rocks: 1) active Sevier-style thrust activity; 2) postorogenic uplift and erosion of the thrust belt, and; 3) active Laramide-style folding.

STRATIGRAPHY

Late Cretaceous to Early Tertiary sedimentary rocks of southwest Utah crop out from the Beaver Dam Mountains to the Table Cliff Plateau (Fig. 1). The stratigraphic units described in this study include the Iron Springs, Kaiparowits, Canaan Peak, Pine Hollow, and Claron formations (Fig. 2). The conglomeratic basal Claron Formation is herein informally named the Formation of Grand Castle because of its stratigraphic position between the Canaan Peak and Pine Hollow formations (Fig. 2). A detailed discussion of the Formation of Grand Castle is provided below.

Iron Springs Formation

The Iron Springs Formation is exposed in the Beaver Dam Mountains and northern Markagunt Plateau (Hintze, 1963; 1980; 1986; 1988) (Fig. 1). The formation consists of 1000 m of sandstone, conglomerate, mudstone, and minor carbonate and coal (Mackin et al., 1976; Mackin and Rowley, 1976; Hintze, 1986; Fillmore, 1989). Grain size within the Iron Springs Formation decreases to the east, away from the Sevier fold and thrust belt (Fillmore, 1989; Fillmore and Middleton, 1989).

The age of Iron Springs Formation ranges from Cenomanian to Campanian. Palynology from the southern Beaver Dam Mountains suggests a Cenomanian to Turonian age for the Iron Springs Formation (Hintze, 1986). An 80 m.y. fission track age was determined from a bentonitic bed below the Iron Springs Formation in the Beaver Dam Mountains (Hintze, 1986).

The Cenomanian to early Campanian age for the Iron Springs Formation reported by Hintze (1986) corresponds with the late Cenomanian to early Campanian ages for the Tropic, Straight Cliffs, and Wahweap formations (Eaton, 1987; Eaton et al., 1987; Eaton and Cifelli, 1988). Although these units are probably time-transgressive, the Iron Springs Formation appears to be correlative with the Tropic, Straight Cliffs, and Wahweap formations in the eastern part of the study area (Bissell, 1952; Threet, 1963; Hintze, 1980).

Kaiparowits Formation

The Kaiparowits Formation is present along the southern Markagunt, Paunsaugunt, Kaiparowits, and Table Cliff plateaus (Gregory, 1950a; 1950b; 1951; Gregory and Moore, 1931; Hackman and Wyant, 1973; Peterson and Kirk, 1977; Sargent and Hansen, 1982). On the Kaiparowits Plateau, the Kaiparowits Formation consist of more than 850 m of fine-grained sandstone and mudstone (Lohrengel, 1969; Eaton et al., 1987). The formation thins to the west (Gregory, 1951) and appears to be absent in the northern Paunsagunt Plateau. In the Cedar Breaks area, sandstones petrographically similar to the Kaiparowits Formation overlie the Iron Springs Formation. Preliminary paleocurrent measurements on planar and trough cross-stratified sandstones in the upper Kaiparowits indicate a northeast-to-eastward flow direction in the area of the Paunsaugunt and Table Cliff plateaus.

The age of the Kaiparowits Formation has been considered both Maastrichtian (Lohrengel, 1969; DeCourten, 1978; DeCourten and Russell, 1985) and Campanian (Bowers, 1972; DeCourten, 1978). Recent faunal studies by Eaton (1987), and Eaton and Cifelli (1988) suggest a middle-to-late Campanian age.

Canaan Peak Formation

The Canaan Peak Formation disconformably overlies the Kaiparowits Formation in the Table Cliff Plateau (Bowers, 1972) and southern Paunsaugunt Plateau (Goldstrand, 1989; 1990b). The westernmost exposures of the Canaan Peak Formation occur on the east side of the Pine Valley Mountains (R. Ernest Anderson, written communication, 1988) where they are preserved in paleovalleys incised into the underlying Navajo Sandstone (Goldstrand, 1990b).

The Canaan Peak Formation consist of boulder to pebble conglomerate, sandstone, and mudstone. At Canaan Peak and on the east side of the Table Cliff Plateau, the formation is approximately 300 m thick (Bowers, 1972). The formation thins or is absent on the west side of the Table Cliff Plateau and along the southern Paunsaugunt Plateau. Imbrication and trough cross-stratification indicate east to northeast paleoflow on the southern Paunsaugunt Plateau (Fig. 3). Along the Table Cliff Plateau, paleocurrents are north to northeast directed (Fig. 3).

Late Campanian pollen were collected from the type-section on the Kaiparowits Plateau (Bowers, 1972) and early Paleocene pollen were collected from the upper part of the Canaan Peak Formation on the southern Table Cliff Plateau (Goldstrand, 1990b). Although the Canaan Peak Formation appears to range in age from late Campanian to early Paleocene, deposition may not have been continuous and intraformational unconformities may exist (Franczyk et al., in press).

Formation of Grand Castle

The Formation of Grand Castle (informal name) is exposed from Parowan Gap to the Table Cliff Plateau. Throughout the study area this thick sequence of conglomerate has been grouped into the basal Claron Formation (Reeside and Bassler, 1922; Thomas and Taylor, 1946; Gregory, 1950a; 1950b; 1951; Bissell, 1952; Cook, 1957; 1960). Along the southern edge of the Table Cliff Plateau, the Formation of Grand Castle has been identified (Goldstrand, 1989; 1990b) where it was previously mapped as Canaan Peak Formation by Bowers (1972), Hackman and Wyant (1973), and Sargent and Hansen (1982). At this locality, the Grand Castle is tectonically tilted and unconformably onlapped by both the Pine Hollow and Claron formations. Because the Pine Hollow Formation lies between the Grand Castle and the Claron, the Grand Castle should not be considered a member of the Claron Formation and is herein designated a separate formation. Thus, this thick conglomeratic sequence is herein informally referred to as the Formation of Grand Castle. A formal designation of this formation is in preparation.

In the Parowan Gap area, the Formation of Grand Castle overlies the erosionally truncated easternmost Sevier thrust faults and associated folds within the Iron Springs Formation. At its type-section east of Parowan (Fig. 1), the Formation of Grand Castle disconformably overlies the Iron Springs Formation. In both localities, the Formation of Grand Castle grades upward into red, calcareous sandstones of the Claron Formation.

The Formation of Grand Castle consists of up to 230 m of conglomerate and sandstone with the thickest section occurring east of Parowan (Goldstrand, 1990b). At its type-section (east of Parowan), the Formation of Grand Castle consist of three facies; a lower boulder conglomerate, a middle sandstone, and an upper cobble-boulder conglomerate. Both conglomeratic facies thin and pinch-out to the south; only the middle sandstone facies is present south of Cedar Breaks National Monument. The middle sandstone facies pinchs-out within 10 km south of the monument. In the Parowan Gap area, only the upper conglomerate is present, unconformably overlying the Iron Springs Formation.

The Formation of Grand Castle, at its eastern extent on the Table Cliff Plateau, appears to be early Paleocene in age. Here, the Grand Castle lies between the upper Canaan Peak and lower Pine Hollow formations, from which early Paleocene palynomorphs have been collected. However, the Formation of Grand Castle to the west grades into the red, calcareous Claron Formation and may be as young as late Paleocene.

Clast imbrication in the lower and upper conglomaterate facies indicate an east to south-southeast paleoflow (Fig. 4). Paleocurrent measurements of trough cross-bedding axes indicate a more easterly paleoflow direction for the middle sandstone facies (Fig. 4).

Pine Hollow Formation

The Pine Hollow Formation of Bowers (1972) is restricted to the Table Cliff Plateau and the west side of Johns Valley. This formation consists of up to 120 m of mudstone, sandstone, pebble conglomerate, and minor limestone (Bowers, 1972; Goldstrand, 1990b). The formation coarsens and thins on the east-limb of the Johns Valley anticline (Fig. 5), where an angular discordance of 10 degrees occurs between the Pine Hollow and Grand Castle. The thickest section of Pine Hollow strata is located near the axis of the Table Cliff syncline (Bowers, 1972) (Fig. 5). Along the axis of the syncline the contact between the Pine Hollow and Canaan Peak formations is disconformable.

The age of the basal Pine Hollow Formation is lower

Paleocene based on palynomorphs (Goldstrand, 1990b). Zircons obtained from a bentonitic mudstone provide a middle Eocene fission track age (50 ± 6 Ma) for the upper Pine Hollow Formation (Bart J. Kowallis, written communications, 1990).

Clast and matrix-supported conglomerate and planar cross-stratified sandstone occur at the edge of the Pine Hollow Formation along the limbs of the Johns Valley anticline. Finer-grained sheet sandstone, red mudstone, and limestone occur within the axis of the Table Cliff syncline. Rooted calcrete zones within the red mudstone and associated mudcracks grade laterally into thin, tabular limestones. Paleoflow was from both the west and northeast (Fig. 5).

Claron Formation

The Claron Formation of Leith and Harder (1908) is exposed throughout the study area (Fig. 1). This unit has been mapped variously as the Claron, Wasatch, and Cedar Breaks formations. The use of Wasatch Formation has been questioned (Robison, 1966; Schneider, 1967; Anderson and Rowley, 1975) because of differences in lithology and age from the type Wasatch Formation in northern Utah and Wyoming. The use of Cedar Breaks Formation (Schneider, 1967) has not gained wide acceptance.

The Claron Formation ranges in thickness from 0 to 165 m in the Beaver Dam Mountains (Hintze, 1986) to 0 to 640 m in the Table Cliff Plateau region (Sargent and Hansen, 1982). Bowers (1972) divided the Claron Formation into three informal members;

a lower pink limestone, a middle white limestone, and an upper varigated sandstone. This study concentrates on the lower member which consists of red, calcareous sandstone, calcareous mudstone, limestone, and minor channelized conglomerate. In this paper, the lower member (of Bowers, 1972) is informally referred to as the lower Claron Formation.

The age of the Claron Formation ranges from Paleocene to middle Oligocene (Bowers, 1972; Rowley et al., 1979; Anderson and Kurlick, 1989). Palynomorph samples collected from the east side of the Pine Valley Mountains indicate an upper Paleocene age for the lower Claron Formation (Goldstrand, 1990b). In the Table Cliff Plateau, a 50 ± 6 Ma fission track age was determine 10 meters below the Pine Hollow-Claron contact suggesting a middle Eocene age for the basal Claron in this area. Gastropod fossils (Viviparus, Physa, and Goniobosis) collected in the lower Claron are similar to those reported from the Paleocene to Eocene Flagstaff Formation of central Utah by LaRocque (1960).

Imbrication data from channelized conglomerates within the lower Claron Formation indicate an east to southeast flow directions (Fig. 6). These channels are sinuous and up to 8 meters deep with steep channel walls. Extensive sandy limestone beds are common in the southern part of the study area. Rare foreset bed orientations from these deposits indicate progradation of deltaic sandstones both to the southeast and to the southwest (Fig. 6).

PETROFACIES AND PROVENANCE

Two different petrofacies are recognized in the study area: volcanic-siliciclastic (the Kaiparowits-Canaan Peak formations) and quartzite-carbonate (Iron Springs-Grand Castle formations) petrofacies. These petrofacies can be differentiated from each other by their sandstone and conglomerate compositions. These petrofacies represent changes in provenance and are important indicators of the evolving paleogeography.

Volcanic-siliciclastic petrofacies

The volcanic-siliciclastic petrofacies includes the middle-to-upper Campanian Kaiparowits and upper Campanian to lower Paleocene Canaan Peak formations (Fig. 7a). Although separated by a disconformity, the sandstone fraction of both formations are very similar. These litharenites and feldspathic litharenites (nomenclature of Folk, 1974) differ from the other Upper Cretaceous and Paleogene sandstones in the study area (Fig. 7), primarily in the relative abundance of feldspar and unique lithic components.

Diagnostic lithic grains include volcanic, siliceous argillitic, and rare metamorphic rock fragments. Volcanic lithic grains consist of microlitic and felsitic fragments and flow banded devitrified tuff fragments. Laminated siliceous argillites commonly contain radiolaria. Metamorphic rock fragments include muscovite-quartz schists and phyllites.

Petrographic and geochemical analysis of volcanic clasts from the Canaan Peak Formation indicate that felsic compositions dominate, but intermediate compositions are common. Welded rhyolitic tuff clasts are also common.

Geochemical comparisons between the Canaan Peak volcanic clasts and the Delfonte Volcanics in southeastern California were made the author, using the standard error of the difference and difference between the mean values (Abbott and Smith, 1989) for 9 major and 4 trace elements. There is no significant difference between means, at the 95% confidence level, for these elements. Thus, volcanic clasts within the Canaan Peak Formation and volcanic lithic fragments in the Kaiparowits Formation appear to be derived from the Delfonte Volcanics of southeastern California (Goldstrand, 1990b).

The Delfonte Volcanics (Hewett, 1956; Evans, 1971) are middle-to-late Jurassic in age and are primarily of felsic compositions (Marzolf, 1983; Busby-Spera, 1988; Marlon A. Nance, written communications, 1988). In southeastern California, the Delfonte Volcanics, Jurassic Aztec Sandstone, and metamorphic (schist and gneiss) Precambrian basement are all involved in Sevier thrust faulting (Evans, 1971; Burchfiel and Davis, 1972; Nelson and Burchfiel, 1979) and are a likely source for part of the Canaan Peak sediments.

Other distinctive clast lithologies in the Canaan Peak Formation are black argillite and chert litharenite clasts. Single clasts show a gradation between the argillite and litharenite lithologies indicating they were derived from the same source. Radiolarians are ubiquitous within the black argillite clasts. The chert litharenite clasts contain grains of chert, quartz, plagioclase, mafic volcanic, and sedimentary lithic fragments of sandstone and siltstone.

These lithologies are petrographically identical to argillites and litharenites of the Mississippian Eleana Formation in southern Nevada. The facies in which the black argillite and litharenite lithologies occur are restricted to the Nevada Test Site and Bare Mountain region (Steven P. Nitchman, personal communication, 1989; Nitchman, 1990). Therefore, the black argillite and chert litharenite clasts in the Canaan Peak and radiolarian chert grains in the Kaiparowits Formation appear to be derived from the Eleana Formation in southern Nevada (Goldstrand, 1989, 1990b).

Quartzite clasts and sandstone grains are common in both the Canaan Peak and Kaiparowits formations. Banded quartzite clasts appear to have been derived from the Prospect Mountain Quartzite and its equivalents.

The mixing of the volcanic provenance and the black argillite-chert litharenite provenance occurred southwest of the eastern Pine Valley Mountains. In the southern Spring Mountains of southern Nevada, synorogenic deposits derived from the Delfonte Volcanics are overthrust by the Contact thrust (Carr, 1980; 1983). No black argillite clasts occur in these deposits, suggesting that mixing of the volcanic and siliciclastic provenances occurred between the Spring Mountains of southern Nevada and the westernmost exposures of the Canaan Peak Formation in the Pine Valley Mountains of southwestern Utah.

In northeastern Utah, the Farrer and Tuscher formations form a coarsening-upward sequence and have sandstone compositions similar to the volcanic-siliciclastic petrofacies. The Farrer and Tuscher formations are coeval, and may be distal equivalents of the Kaiparowits and Canaan Peak formations (Lawton, 1983; Franczyk et al., in press; Franczyk et al., in press).

Quartzite-carbonate Petrofacies

The Upper Cretaceous Iron Spring Formation and the lower Paleocene Formation of Grand Castle form another coarsening-upward sequence that represent the quartzite-carbonate petrofacies. This petrofacies lies to the north and northwest of the volcanic-siliciclastic petrofacies, and indicates a distinctly different provenance.

The sandstone petrology of the Iron Springs and Grand Castle are similar (Fig. 7b). Sandstone compositions are sublitharenites to litharenites (nomenclature of Folk, 1974), with the major lithic component being carbonate and silicified limestone grains. Silicified limestone grains were petrographically differentiated from chert grains by the presences of dolomite crystals, fossil fragments (other than radiolarians), or zones of carbonate preserved in a polycrystalline-quartz matrix.

Clast compositions in the Formation of Grand Castle include quartzite, limestone, silicified limestone, and minor dolostone. Quartzite clasts are largely derived from the Prospect Mountain

Quartzite. Upper Paleozoic fossils are common in the silicified limestone clasts and include the sponge Chaetetes, fusulinid foraminifera, rugose corals, bryozoans, crinoid columns, and brachiopods.

These clast lithologies, when combined with the south-southeast to east paleocurrent directions, indicate that the Formation of Grand Castle was derived from the Wah Wah and Blue Mountain thrust sheets to the west (Goldstrand, 1989). The Prospect Mountain Quartzite is exposed in the upper plate of the Wah Wah thrust, whereas Paleozoic limestone and dolomite strata are exposed in the upper plate of the Blue Mountain thrust (Miller, 1963; 1966). Fillmore (1989) proposed that the provenance for the Iron Spring Formation was also the Blue Mountain and Wah Wah thrust sheets. Thus, the Iron Spring Formation and Formation of Grand Castle were both derived from these thrust sheets.

The presence of the middle sandstone facies between conglomeratic facies of the Formation of Grand Castle appears to be related to a change in provenance rather than tectonism. The petrology of these quartzarenites and sublitharenites differ slightly from other sandstones of the Formation of Grand Castle. Within the middle sandstone facies, fine to very-fine monocrystalline quartz with abundant quartz overgrowths and dust-rims dominate. These petrographic attributes are similar to the Navajo Sandstone which is exposed in the lower plate of the Blue Mountain thurst and may have been the source for the middle sandstone facies of the Formation of Grand Castle.

PALEOGEOGRAPHY

Figures 8 to 14 represent interpretations of the paleogeography from the early Campanian to the middle Eocene for southwest Utah. The generalized geologic map in Figure 1 was used as the base map for these paleogeographic reconstructions. Inset regional maps (from Stewart, 1980) show major thrust and strike-slip faults in southeastern California, southern Nevada, and southwestern Utah. The regional maps have not been corrected for Basin and Range extension. Restorations of large scale Tertiary extension (Wernicke, et al., 1988; Levy and Christie-Blick, 1989) bring the proposed source areas closer to the depositional basins in southwestern Utah, but do not substantially effect the paleogeographic reconstructions.

Paleogeographic interpretations are speculative and highly generalized. These reconstructions are based on: 1) outcrop data (Fig. 1); 2) limited well data (Doelling and Davis, 1978); 3) paleocurrent data; 4) facies analysis; 5) changes in provenance; and 6) location of unconformities and the bounding strata. Relief of highlands are qualitative, being based on grain size and provenance.

Early Campanian

During the early Campanian, sandy braided fluvial systems of the Iron Springs Formation drained the active Sevier thrust belt

(Fig. 8). Detritus was shed off the Proterozoic and Paleozoic strata exposed in the Wah Wah and Blue Mountain thrust sheets. Precambrian quartzites in the upper plate of the Wah Wah thrust (Miller, 1963; 1966) may have been passively uplifted during movement along the structurally lower Blue Mountain thrust. Antecedent drainages allowed mixing of different clast lithologies from the two thrust sheets. Similar relationships have been documented in the Paris-Willard and Absaroka thrusts of northern Utah (Steidtmann and Schmitt, 1988; Schmitt and Steidtmann, 1990).

Depositional environments in the upper Iron Springs Formation include alluvial fan to fluvial braidplain settings (Fillmore, 1989). The northeast paleocurrent directions in the Iron Springs (Fillmore, 1989) suggest thrust-parallel flow in the proximal foreland basin. Paleocurrent directions in the more distal foreland basin indicate an eastward flow toward the retreating Cretaceous seaway.

Middle-to-Late Campanian

A change in the provenance from Paleozoic strata exposed in the west, to a volcanic and siliciclastic source derived from the southwest occurred during the middle-to-late Campanian (Fig. 9). The middle-to-late Campanian Kaiparowits Formation (Eaton and Cifelli, 1988) had a provenance in southeastern California and southern Nevada. As previously discussed, the source for the volcanic component was the Delfonte Volcanics in southern Nevada and southeastern California (Fig. 9). The source for the black, radiolarian argillites was the Mississippian Eleana Formation in southern Nevada (Fig. 9).

The Eleana lithologies were probably transported through an antecedent drainage traversing the Sevier hinterland and thrust belt, similar to that proposed by Burbank and Raynolds (1988) for the Himalayan thrust belt. Mixing of the volcanic and siliciclastic provenances occurred in the foreland basin, between the Spring Mountains of southern Nevada and the Pine Valley Mountains of southwestern Utah.

A longitudinal trunk stream flowed northeast, parallel to the Sevier highland (Fig. 9). The Mogollon Highland of central and southwest Arizona had developed prior to the Late Cretaceous (Dickinson, 1981; Bilodeau, 1986; Eaton et al., 1987) and may have impeded flow to the east and south (Molenaar, 1983).

In southwest Utah, the Kaiparowits Formation was deposited in an east to northeast flowing meandering river system. Along the northern Markagunt and Paunsaugunt plateaus the Kaiparowits Formation thins, is generally coarser, and unconformably overlies the Iron Springs or Wahweap formations. These relationships suggest that the northern parts of these plateaus were topographically positive during the middle-to-late Campanian, and formed the northern boundary of Kaiparowits deposition. The thick section of Kaiparowits strata and northeasterly change in paleocurrent directions in the Table Cliff Plateau region may indicate the beginning of Laramide-style folding of the Sevier foreland basin at this time.

A thrust fault and associated fault propogation fold deforms the Iron Springs Formation in the Parowan Gap area. This thrust represents the eastern extent of the Sevier thrust system and clearly deforms foreland basin deposits. The Parowan Gap thrust is on trend with and exhibits a similar fault geometry as the Iron Springs thrust (Mackin, 1960; Mackin and Rowley, 1976; Mackin et al., 1976; Van Kooten, 1988). Because this fault cuts latest Cretaceous strata and is unconformably overlain by the early Paleocene Formation of Grand Castle, the timing of latest Sevier-style deformation is constrained between latest Cretaceous and earliest Paleocene time.

Late Campanian

During the late Campanian, and deposition of the lower Canaan Peak Formation, the regional drainage was similar to that of the middle-to-late Campanian (Fig. 10). The Canaan Peak strata represents coarse, braided fluvial deposition (Jones, 1989; Schmitt et al., in press). Paleocurrent and petrographic data suggest that the Kaiparowits and Canaan Peak river systems formed a through-going drainage to northern Utah and are the proximal deposits of the Tuscher and Farrer formations. Laramide deformation related to the Circle Cliffs Uplift may be recorded in paleocurrent divergence from east to northeast in the southern Paunsaugunt Plateau (Goldstrand, 1989).

The Kaiparowits and Canaan Peak formations have the same provenance and are a coarsening-upward sequence that may record cessation of Sevier-style deformation. Imposing the model of Heller et al. (1988) the fine-grained Kaiparowits Formation was deposited during active thrusting, whereas the coarse-grained deposits of the Canaan Peak Formation may represent postorogenic erosion and rebound of thrust sheets in southeastern California and southern Nevada.

Thinning of the Canaan Peak Formation on the northern Paunsaugunt Plateau is partially a result of later uplift and erosion during the early Paleocene. However, incision into the underlying Kaiparowits Formation and fining of the Canaan Peak Formation on the southern Paunsaugunt Plateau suggest that the northern plateau region was topographically positive during the late Campanian.

Reworked Canaan Peak conglomerate and sandstone are present within deltaic deposits of the Claron Formation along the southern Markagunt Plateau. Foreset bedding orientations indicate a southeast paleoflow (Fig. 6). Thus, the Canaan Peak Formation may have been deposited to the northwest of the southern Markagunt Plateau and was later eroded during the late Paleocene or Eocene.

In the eastern Pine Valley Mountains, the Canaan Peak Formation unconformably overlies the Jurassic Navajo Sandstone. However, a few kilometers to the west of this exposure, 1,250 m of Late Cretaceous rocks are present in the Pine Valley Mountains (Cook, 1957; 1960). To the east, on the southern Paunsaugunt

Plateau, the Canaan Peak Formation was deposited over middle-to-late Campanian strata. These stratigraphic and structural relationships (Cook, 1957) suggest that a Late Cretaceous monocline developed in the vicinity of the Hurricane Fault. Development of this monocline resulted in uplift of Jurassic rocks prior to the deposition of the Canaan Peak Formation, but after deposition of the Campanian Iron Springs and Kaiparowits formations. Young (1979) suggests initial monoclinal folding of the Hurricane structure in northern Arizona occurred in the Late Cretaceous, which agrees with the timing of folding along the eastern Pine Valley Mountains.

Maastrichtian

No Maastrichtian age rocks have been identified in southwest Utah. Late Campanian palynomorphs have been recovered from the lower Canaan Peak Formation (Bowers, 1972). Early Paleocene palynomorphs have been collected form the uppermost Canaan Peak Formation by Goldstrand (1990b). The lack of known Maastrichtian strata may indicate a period of nondeposition (Fig. 11).

In central Utah, nondepostion occurred during the latest Campanian or early Maastrichtian time as a result of the development of the San Rafael uplift (Cross, 1986; Lawton, 1986; Franczyk and Nichols, 1988; Franczyk et al., in press). A broad, low amplitude uplift (a precursor to Laramide tectonism) may be responsible for Maastrichtian depositional hiatus in southwest Utah.

Early Early Paleocene

During the early early Paleocene (Fig. 12) the uppermost Canaan Peak Formation consisted of a east to north-northeast flowing braided river system. Although the southern extent of the Canaan Peak fluvial system is unknown, the Kaibab uplift had developed by this time (Cross, 1986) and may have formed the southern boundary.

Along the southern Table Cliff Plateau, conglomerates of the Formation of Grand Castle overlie and rework the Canaan Peak Formation (Goldstrand, 1989; 1990b). Architectural-element analysis (methods of Miall, 1985; 1988) of the Formation of Grand Castle indicates that the upper and lower conglomerate facies represent gravelly, braided fluvial deposition. The middle sandstone facies was deposited in a sandy, braided river environment.

In the Table Cliff Plateau region, paleocurrents in the Formation of Grand Castle are east-directed. The main channel connecting the western and eastern exposures of the Formation of Grand Castle projects beneath younger volcanic rocks north of the Markagunt and Paunsaugunt plateaus.

The Formation of Grand Castle is absent on the Paunsaugunt and southern Markagunt plateaus, indicating this region was topographically positive. Isopach mapping of the Formation of Grand Castle in the Markagunt Plateau and Parowan Gap region indicate that the depocenter is east of Parowan (Goldstrand, 1990b). The southern basin margin is located about 10 km south of

Cedar Breaks National Monument. The northern basin margin has not been identified due to volcanic cover and disruption by the Hurricane Fault, but isopach mapping suggests that the Formation of Grand Castle thins to the north of Parowan. In the Parowan Gap area, the lower conglomerate and middle sandstone facies of the Formation of Grand Castle are absent (Goldstrand, 1990b). A topographic high at Parowan Gap may have existed until deposition of the upper conglomerate of the Formation of Grand Castle.

The Iron Springs Formation and Formation of Grand Castle represent a coarsening upward sequence consisting of sediment derived from the Wah Wah and Blue Mountain thrusts (Fillmore, 1989; Goldstrand, 1989). Coarse clastics of the Formation of Grand Castle may represent the postorogenic uplift of the thrust belt.

If the Canaan Peak Formation and Formation of Grand Castle both represent postorogenic uplift of the Sevier thrust belt, the southern part of the belt was uplifted during the Late Cretaceous to early Paleocene. The northern parts of the thrust belt (in southwest Utah) were uplifted later, during the early Paleocene.

Late Early Paleocene

Late early Paleocene saw the partitioning of the foreland basin and development of internally drained, intermontane basins (Fig. 13). Late early Paleocene is considered to be a time when Laramide-style deformation was at a maximum in southwest Utah.

The gradational contact between the Formation of Grand Castle and the Claron Formation in the northern Markagunt Plateau suggest continuous deposition into the late Paleocene in this area. Much of the Pine Valley Mountains, the Markagunt, and the Paunsaugunt plateaus were topographically positive areas. Although the northern Paunsaugunt Plateau may have been affected by Late Cretaceous anticlinal folding, upwarping and erosion appears to have been reactivated in the early Paleocene. This northeast trending anticlinal structure appears to project onto the southern Markagunt Plateau, and formed the southern boundary of the Grand Castle basin.

This anticlinal flexure may have controlled development of later structures. The axial trend of this anticline may have formed a north to northwest declivity on which the Eocene age Rubys Inn thrust (Lundin, 1989; Davis, 1990) (see Fig. 3) ramped up to the south and southeast through the Claron Formation.

Deposition of the Pine Hollow Formation was controlled by the development of the Johns Valley anticline and folds to the east (the Circle Cliffs uplift, Upper Valley and Escalante anticlines). Paleocurrents in the Pine Hollow Formation diverge from these upwarps. Recycled conglomerates from both the Canaan Peak Formation and the Formation of Grand Castle form small alluvial fans and debris flow deposits on the east limb of the Johns Valley anticline. Fluvial and sheet-flood sandstones grade laterally into playa-mudflat and lacustrine deposits within the basin center.

On the east side of the Table Cliff Plateau, coarse fluvial conglomerates and sandstones of the Pine Hollow Formation overlie conglomerates of the Canaan Peak with an angular discordance of approximately 10 degrees. Red siliceous siltstone clasts in the basal conglomerate of the Pine Hollow indicate a partial source from the Brushy Basin Member of the Morrison Formation exposed on the west flank of the Circle Cliffs uplift near Escalante.

In the northern Table Cliff Plateau an increase in clast and sandstone maturity has been recognized in the uppermost Canaan Peak (Goldstrand, 1990b). Goldstrand (1990b) interpreted this increase in maturity as recycling and redeposition of the Canaan Peak conglomerates during initial Laramide folding before deposition of the Pine Hollow Formation.

The late early Paleocene sediments of the lower Pine Hollow Formation have red overbank fines, pedogenic calcrete zones with deep tap-rootlets, mudcracks, and rare gypsum. These sediments suggest a semi-arid environment dominated by periods of dessication.

Late Paleocene to middle Eocene

Beginning in the late Paleocene (Fig. 14) and continuing into the Eocene, a slowly subsiding basin developed in southwest Utah. In the Table Cliff Plateau, the Pine Hollow Formation grades upward into the lower Claron Formation. The Claron Formation overlaps structural highs of the Johns Valley anticline, indicating that Laramide deformation had ceased prior to Claron deposition in this region.

Around these overlap regions, nearshore lacustrine facies occur and may have developed around small islands during the initial phase of lake development. Cross-stratified sandstones, symmetric ripples, oncolites, stromatolites, charophytes, bivalves, and gastropods are common in the nearshore lacustrine facies. Other nearshore lacustrine facies occur along the southern Paunsaugunt Plateau within the basal Claron Formation. Stromatolites, oncolites, gastropods, bivalves, and reworked Canaan Peak conglomerates occur in these shore facies.

Laterally extensive floodplain deposits with strong pedogenic overprinting (Mullet et al., 1988; Mullet, 1989) occur in the northern Markagunt and Paunsaugunt plateaus. Deeply incised pebble conglomerate channels within these paleosols may be a result of a lowering of base-levels related to lake level fluctuations. Pebble lithologies suggest that the upper Formation of Grand Castle was the source of these channel conglomerates.

The basal Claron is time-transgressive, ranging in age from late Paleocene in the Pine Valley Mountains to middle Eocene in the Table Cliff Plateau region. Facies relationships suggest that during the late Paleocene a lake system began in the west (on the east-side of the Pine Valley Mountains) and along the southern Markagunt and Paunsaugunt plateaus. During the late Paleocene to middle Eocene, in the Table Cliff Plateau, intermontane playa-mudflat and lacustrine deposition was occurring in the Pine Hollow basin.

The Pine Hollow basin and its structural boundaries were not overlapped by lacustrine deposits of the Claron Formation until the middle Eocene. This overlap relationship indicates that the

Table Cliff region was the eastern boundary of the Claron basin from late Paleocene to early Eocene. A Gilbert-style delta prograded southwestward into lake deposits at the base of the Claron Formation on the southern Table Cliff Plateau suggesting the eastern basin margin was still near this region during the middle Eocene. In the Beaver Dam Mountains, alluvial sediments derived off the Navajo Sandstone are interbedded with lacustrine mudstone and limestone of the Claron Formation suggesting this was the western boundary of the Claron basin.

This lake was initially relatively shallow (a minimum of 4 m deep based on deltaic foreset thicknesses) and was boardered to the west and east by uplands and bounded to the north by a relatively flat floodplain. Along the southern Markagunt Plateau another Gilbert-type delta is exposed 170 m above the Claron-Kaiparowits unconformity. Foreset thicknesses suggest the lake was a minimum of 12 m deep at this time.

The thick sequences of calcrete paleosols in the lower Claron Formation suggest slow subsidance (Bown and Kraus, 1981; Shuster and Steidtmann, 1987) over an extended period of time (Retallack, 1983; Reading, 1986). Rootlets penetrate downward 1 to 1.5 meters, indicating well-drained paleosols and a low water table (Retallack, 1988). These calcrete paleosols suggest that semi-arid climatic conditions (Reading, 1986) persisted during Claron deposition (Late Paleocene to Eocene). As noted above, the Pine Hollow Formation appears to document semi-arid climatic conditions by late early Paleocene.

The interpretation of an apparent change to a semi-arid climate in southwest Utah during the Paleocene to Eocene differs from the work of Kraus (1984), Wolfe and Upchurch (1987), and Kraus and Brown (1988) who proposed a warm, humid climate during the Paleocene to early Eocene throughout the Western Interior. The disparity in the paleoclimate of southwest Utah with that of the rest of the Western Interior is probably related to orographic rainshadow effects within these western intermontane basins.

SUMMARY

During the Late Cretaceous, Sevier-style tectonism in southeastern California, southern Nevada, and western Utah formed highlands from which sediment was shed eastward into southwestern Utah. Beginning in the latest Cretaceous and throughout the Paleogene, Laramide-style tectonism resulted in the partitioning of the foreland basin into smaller intermontane basins.

During the early Campanian, braided fluvial systems transported sediment east and northeast from the Wah Wah and Blue Mountain thrust sheets of southwestern Utah. Sandstones of the Iron Springs Formation are folded and faulted by the easternmost thrust faults of the Sevier thrust belt (Parowan Gap-Iron Springs thrust), indicating Sevier deformation continued into the latest Cretaceous.

Another fluvial system developed in the middle-to-late Campanian, depositing sandstone and mudstone of the Kaiparowits

Formation. These east-to-northeast directed meandering rivers were derived from volcanic, siliciclastic, and metamorphic sources in southeastern California and southern Nevada (the Jurassic Delfonte Volcanics, Mississippian Eleana Formation, and Precambrian basement).

Throughout the late Campanian to early Paleocene the conglomeratic Canaan Peak Formation was deposited. These braided river deposits have the same source as the Kaiparowits Formation. The Canaan Peak fluvial system was structurally controlled between a topographic high on the northern Paunsaugunt Plateau and uplifts to the east. Divergence in paleoflow, from east to north-northeast, may signal the initiation of Laramide tectonism during the late Campanian. No sediments of Maastrichtian age have been reported from southwestern Utah which may indicate a period of nondeposition related to Laramide deformation.

An east to south-southeast flowing braided fluvial system (Formation of Grand Castle) formed during the early Paleocene. Like the Iron Springs Formation, clastics of the Formation of Grand Castle were derived from the Wah Wah and Blue Mountain thrust sheets.

The Iron Springs and Kaiparowits formations appear to represent synorogenic sedimentation derived from active Sevier thrusting during the Campanian. However, the laterally extensive conglomerates of the Canaan Peak Formation and Formation of Grand Castle may represent a northward propagation of postorogenic erosion and rebound of the Sevier fold and thrust belt thoughout the late Campanian to early Paleocene.

The Pine Hollow Formation provides evidence of Laramide partitioning of the Sevier foreland basin during the early Paleocene to middle Eocene. This unit represents an internally drained basin that received sediment from both the west and northeast during the development of Laramide folds.

The Claron Formation is a time-transgressive sequence of fluvial, deltaic, and lacustrine rocks. Lacustrine deposition may have began in the west and south and transgressed north and northeast over relatively flat floodplain deposits. In the Table Cliff Plateau region the lower Claron Formation did not overlap the Pine Hollow intermontane basin until middle Eocene time. Overlap of the anticlinal structures by the Claron suggest cessation of Laramide-style deformation occurred around 50 Ma.

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Fig. 3: Paleocurrent data from the Canaan Peak Formation (stippled). I= imbrication measurements, T= mean of trough axes.

Fig. 4: Paleocurrent data from the Formation of Grand Castle (stippled). I= imbrication measurements, T= mean of trough axes.

Fig. 5: Generalized geologic map of the Table Cliff Plateau (from Goldstrand, 1990b) and paleocurrent data for the Pine Hollow Formation (stippled). I= imbication measurements, P= planar cross-stratification measurements. Kk= Kaiparowits Formation, TKcp= Canaan Peak Formation, Tgc= Formation of Grand Castle , Tc= Claron Formation.

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Fig. 14: Paleogeographic reconstruction during the late Paleocene to middle Eocene (deposition of the upper Pine Hollow and lower Claron formations). Arrows indicate general paleocurrent directions.

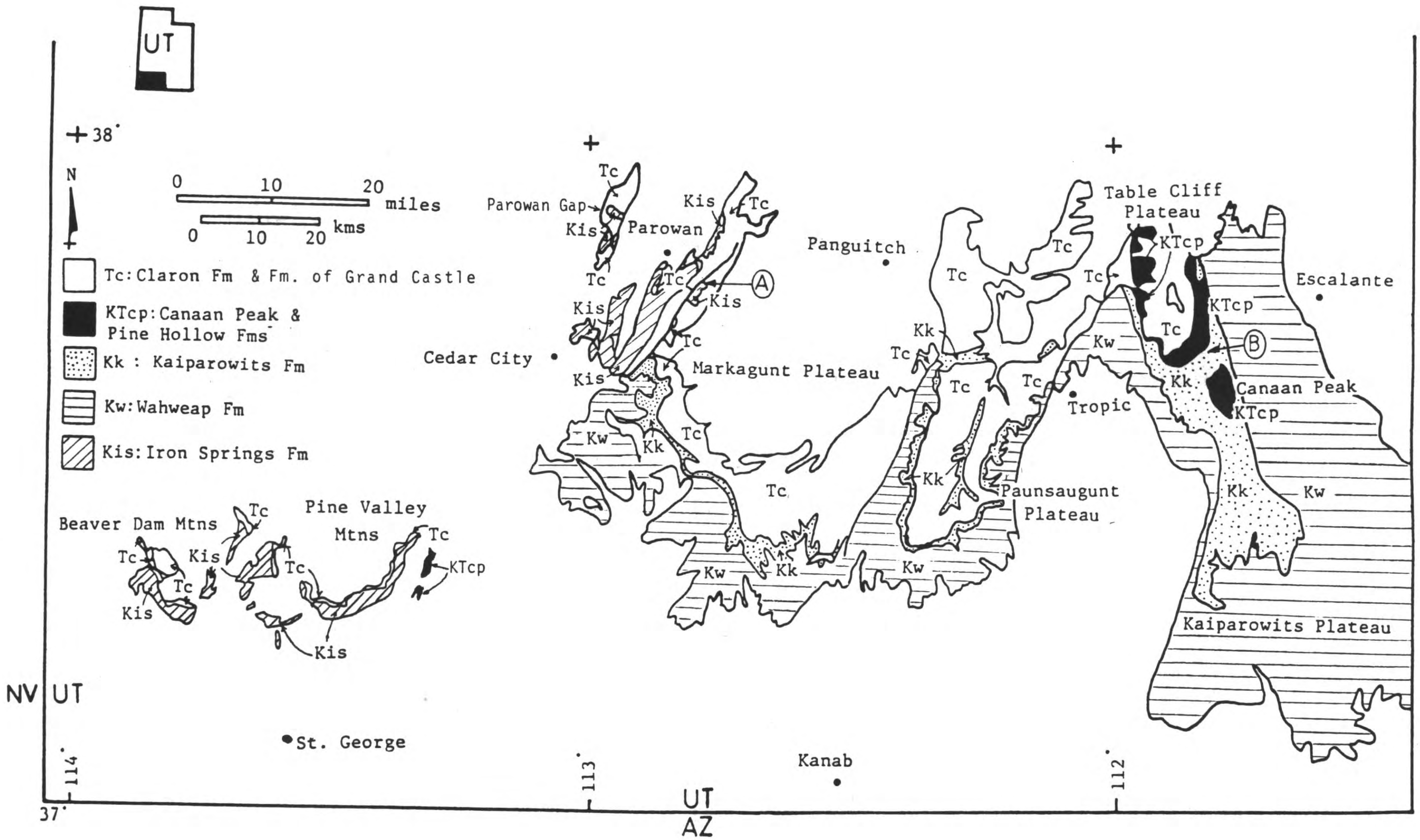
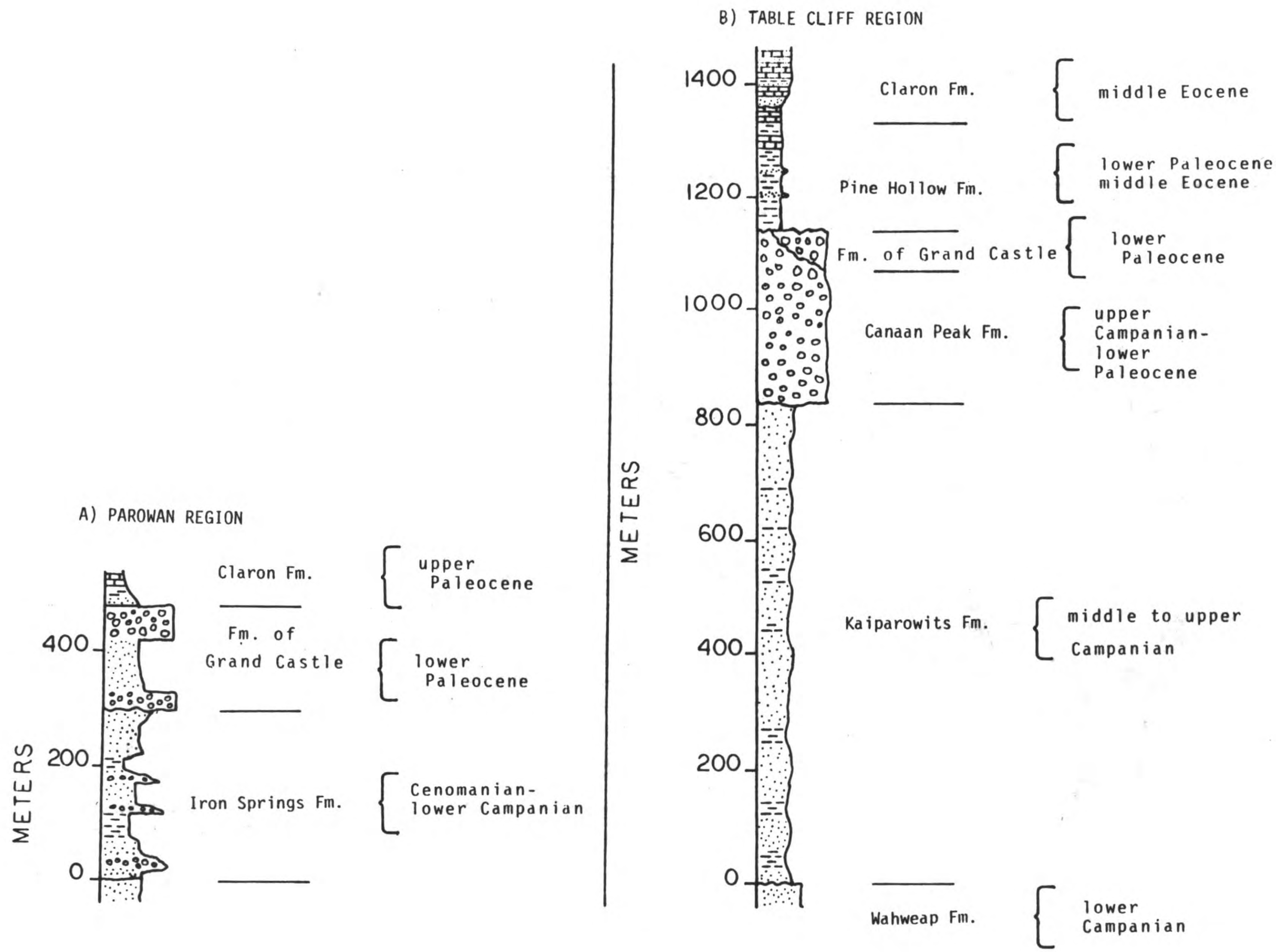

UT
+38°
N
0 10 20 miles
0 10 20 kms
Tc: Claron Fm & Fm. of Grand Castle
KTcp: Canaan Peak & Pine Hollow Fms
Kk : Kaiparowits Fm
Kw: Wahweap Fm
Kis: Iron Springs Fm
Parowan Gap
Parowan
Cedar City
Markagunt Plateau
Panguitch
Table Cliff Plateau
Escalante
Tropic
Canaan Peak
Paunsaugunt Plateau
Kaiparowits Plateau
Beaver Dam Mtns
Pine Valley Mtns
St. George
Kanab
A
B
Tc
Kis
Kk
Kw
KTcp
NV UT
UT AZ
114°
113°
112°
37°

Figure 1.



B) TABLE CLIFF REGION
1400
1200
1000
800
600
400
200
0
METERS
Claron Fm.
middle Eocene
Pine Hollow Fm.
lower Paleocene-
middle Eocene
Fm. of Grand Castle
lower
Paleocene
Canaan Peak Fm.
upper
Campanian-
lower
Paleocene
Kaiparowits Fm.
middle to upper
Campanian
Wahweap Fm.
lower
Campanian
A) PAROWAN REGION
400
200
0
METERS
Claron Fm.
upper
Paleocene
Fm. of
Grand Castle
lower
Paleocene
Iron Springs Fm.
Cenomanian-
lower Campanian

Figure 2.

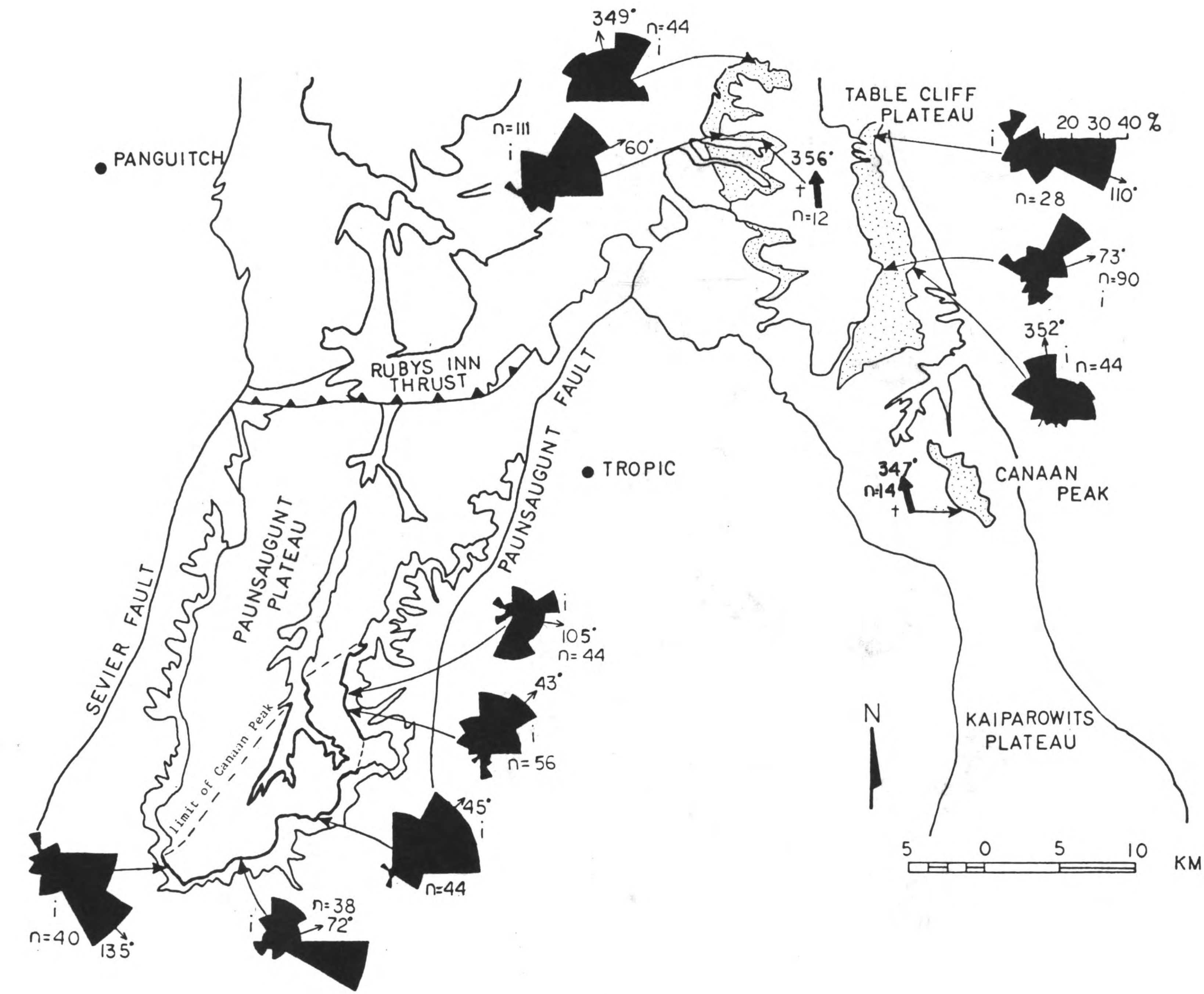
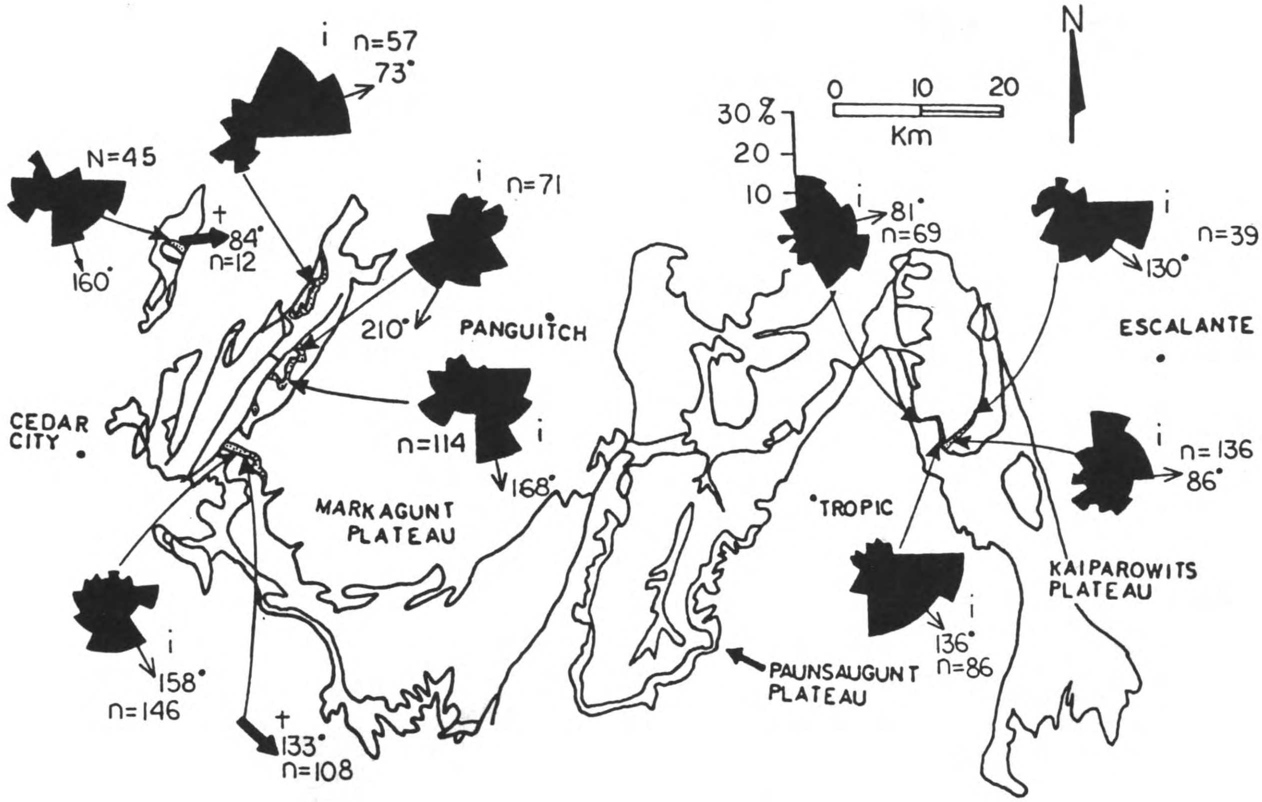

349°
n=44
n=111
60°
PANGUITCH
TABLE CLIFF
PLATEAU
20 30 40 %
356°
n=12
n=28
110°
73°
n=90
352°
n=44
RUBYS INN
THRUST
PAUNSAUGUNT FAULT
TROPIC
347°
n=14
CANAAN
PEAK
SEVIER FAULT
PAUNSAUGUNT
PLATEAU
105°
n= 44
43°
n= 56
limit of Canaan Peak
N
KAIPAROWITS
PLATEAU
45°
n=44
5 0 5 10 KM
n=40
135°
n= 38
72°

Figure 3.



N
0 10 20
Km
30 %
20
10
i n=57
73°
N=45
+
84°
n=12
160°
i n=71
210°
PANGUITCH
i
81°
n=69
i
n=39
130°
ESCALANTE
CEDAR CITY
n=114
i
168°
i n=136
86°
MARKAGUNT PLATEAU
TROPIC
KAIPAROWITS PLATEAU
i
158°
n=146
+
133°
n=108
PAUNSAUGUNT PLATEAU
i
136°
n=86

Figure 4.

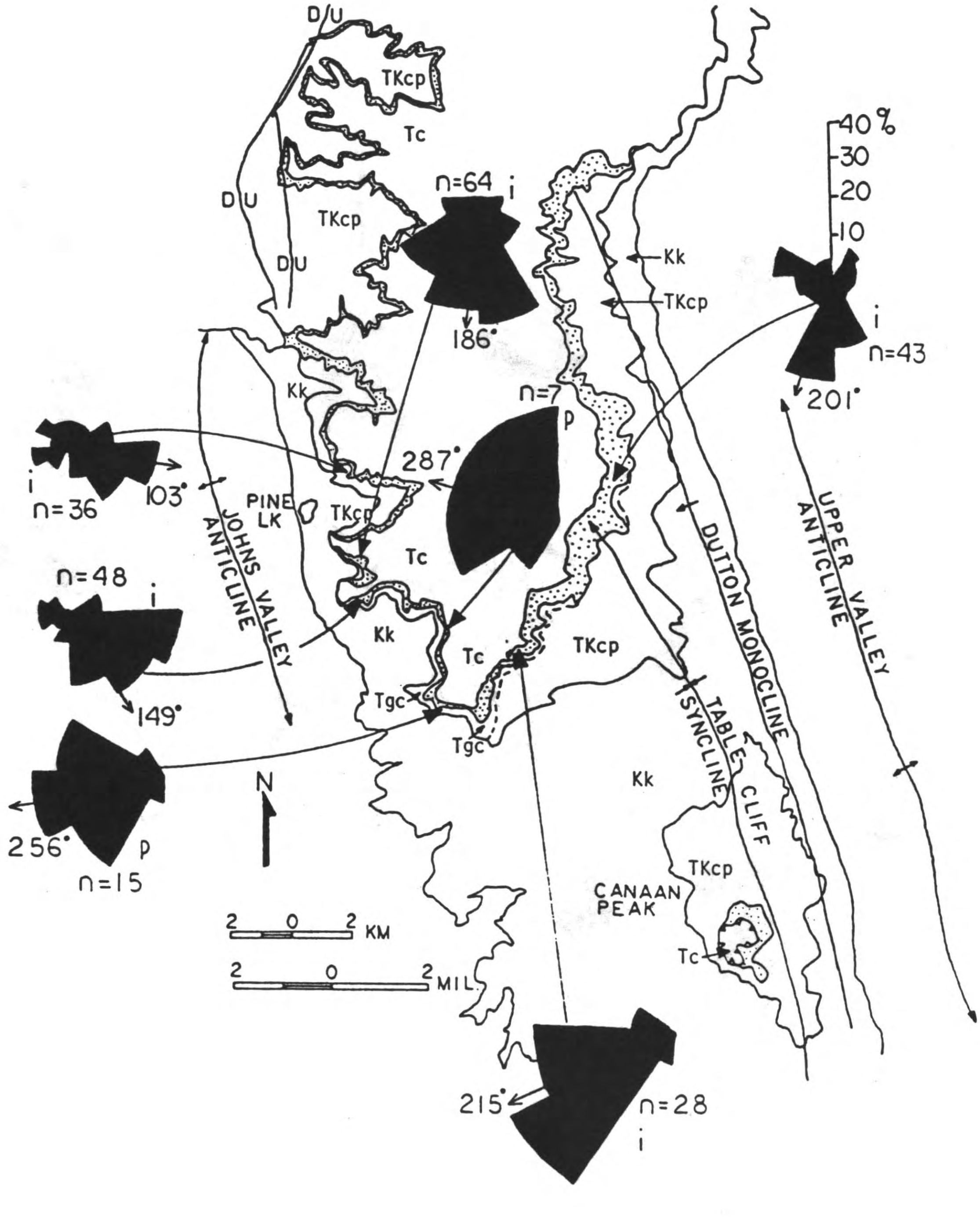

D U
TKcp
Tc
D U
TKcp
D U
n=64 i
186°
Kk
TKcp
40 %
30
20
10
i
n=43
201°
n=7
p
287°
i
n=36
103°
PINE LK
TKcp
JOHNS VALLEY ANTICLINE
Tc
n=48
i
149°
Kk
Tc
TKcp
Tgc
Tgc
DUTTON MONOCLINE
UPPER VALLEY ANTICLINE
TABLE CLIFF SYNCLINE
N
256°
p
n=15
Kk
TKcp
CANAAN PEAK
Tc
2 0 2 KM
2 0 2 MIL
215°
n=28
i

Figure 5.

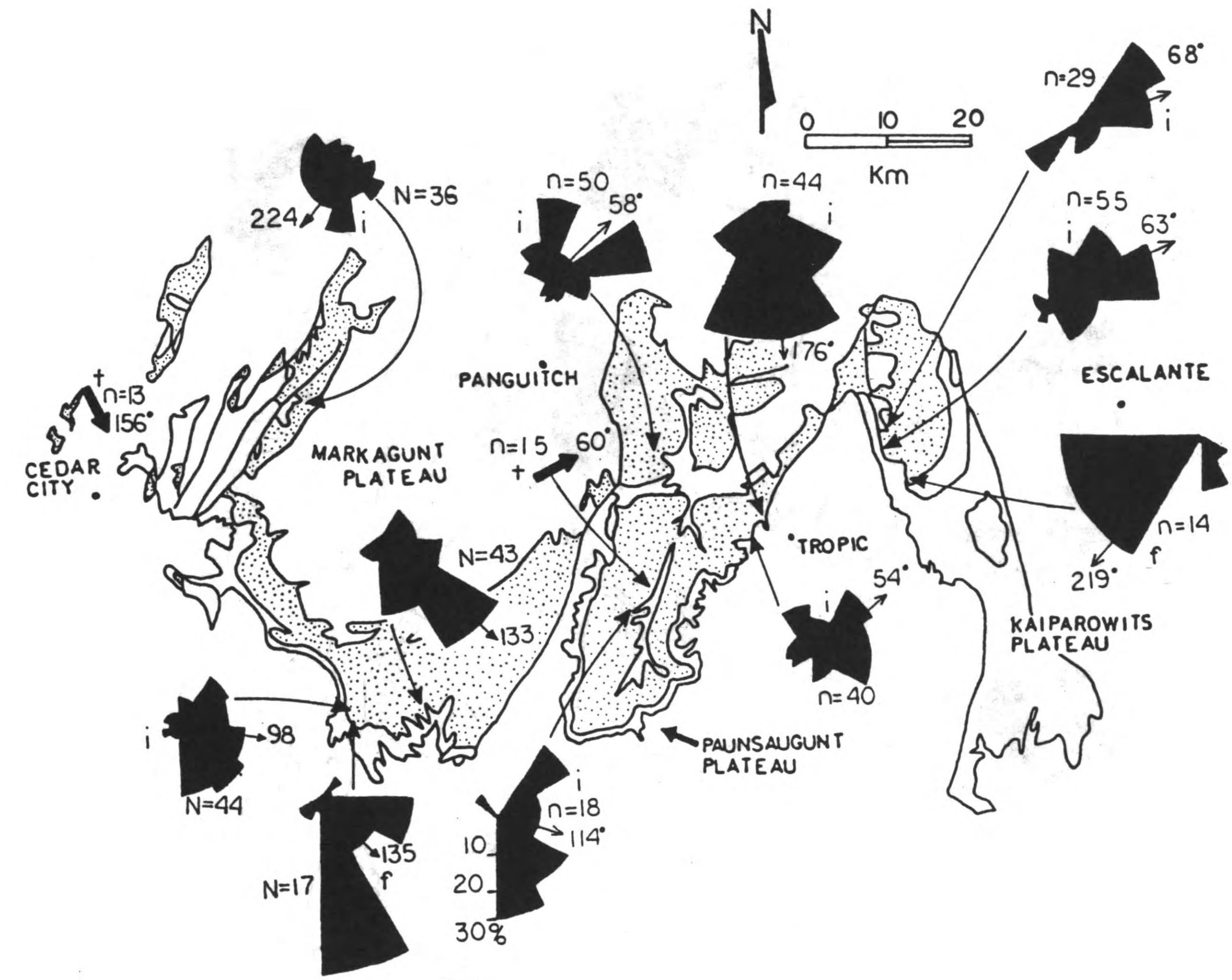
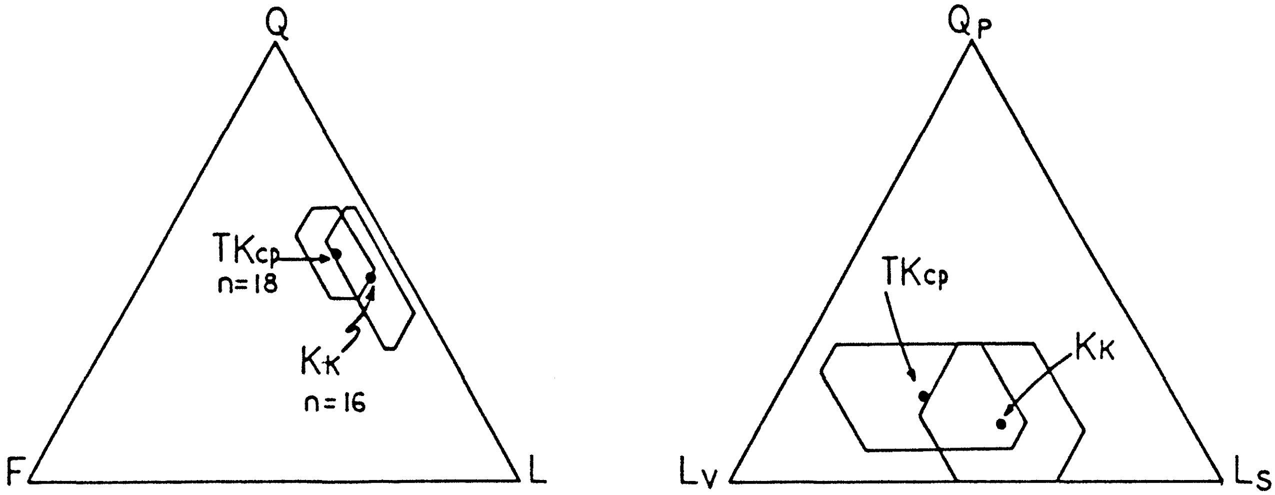
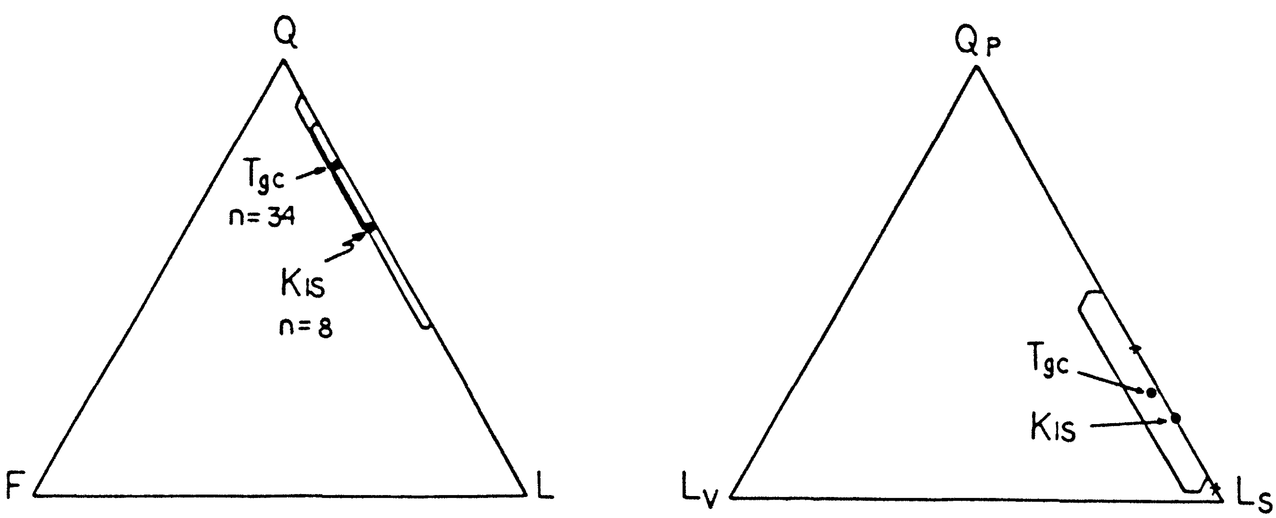

N
0
10
20
Km
n=29
68°
i
n=55
63°
i
ESCALANTE
n=14
f
219°
KAIPAROWITS
PLATEAU
n=44
i
176°
n=50
58°
i
224
i
N=36
n=13
156°
CEDAR
CITY
MARKAGUNT
PLATEAU
PANGUITCH
n=15
60°
N=43
133
TROPIC
54°
i
n=40
PAUNSAUGUNT
PLATEAU
i
98
N=44
N=17
135
f
i
n=18
114°
10
20
30%

Figure 6.



CANAAN PEAK FORMATION (TKcp) & KAIPAROWITS FORMATION (Kk)



FORMATION OF GRAND CASTLE (Tgc) & IRON SPRINGS FORMATION (Kis)

Figure 7.

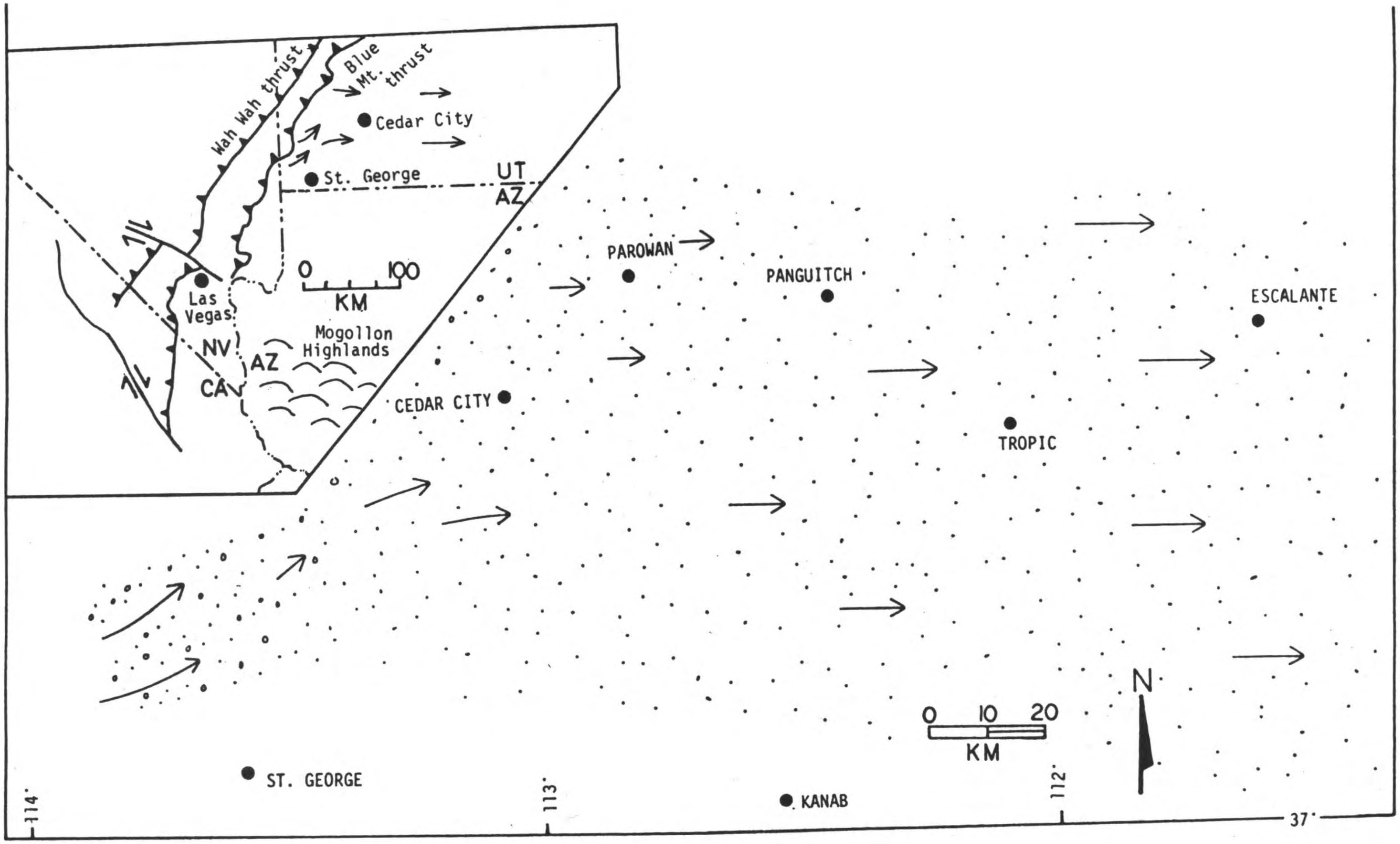

Wah Wah thrust
Blue Mt. thrust
Cedar City
St. George
UT
AZ
0 100
KM
Las Vegas
NV
AZ
CA
Mogollon Highlands
PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
0 10 20
KM
N
ST. GEORGE
KANAB
114°
113°
112°
37°

Figure 8.

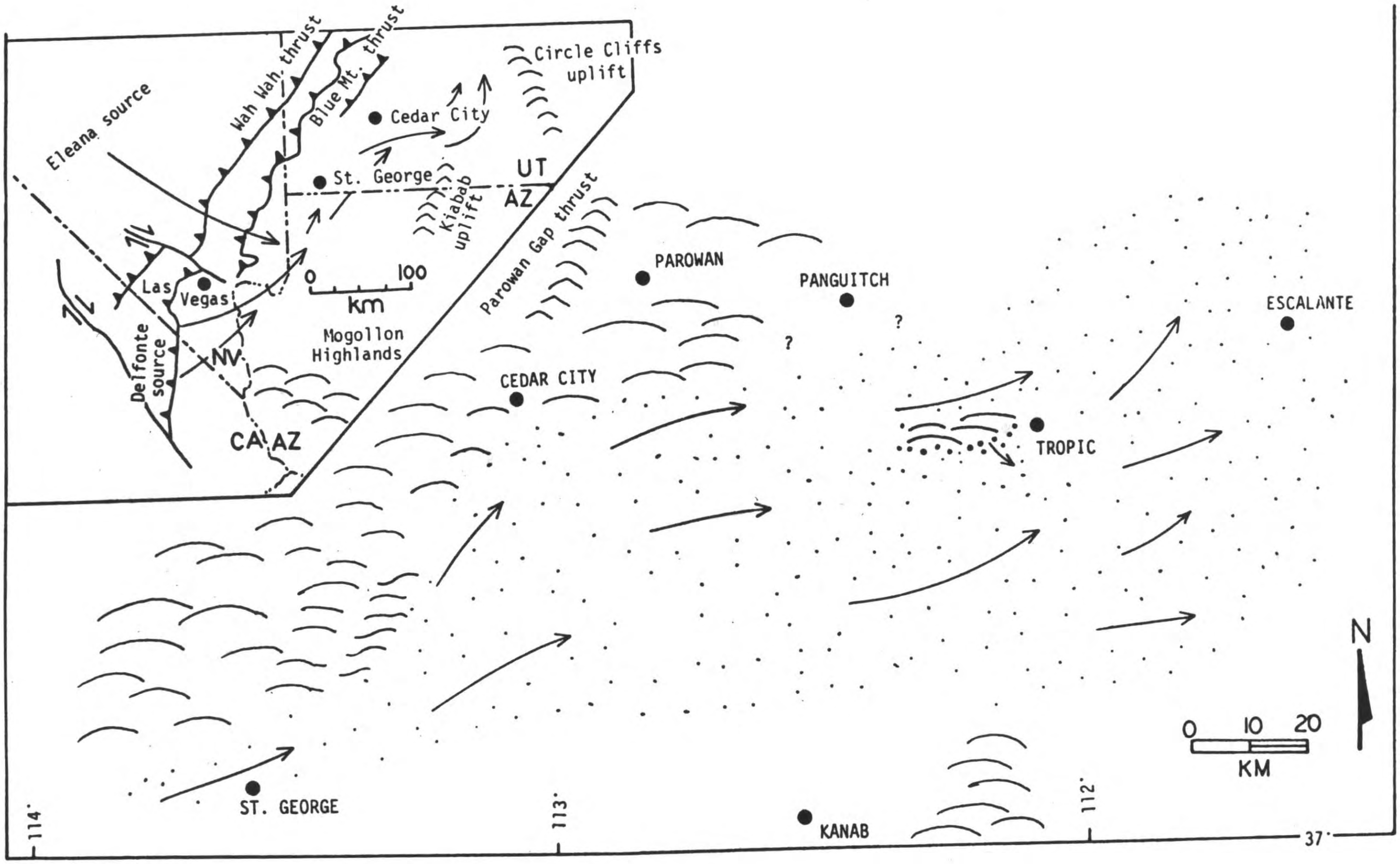

Circle Cliffs
uplift
Wah Wah thrust
Blue Mt. thrust
Eleana source
Cedar City
St. George
UT
AZ
Kiabab
uplift
Parowan Gap thrust
Las
Vegas
0
100
km
Mogollon
Highlands
Delfonte
source
NV
CA
AZ
PAROWAN
PANGUITCH
ESCALANTE
?
?
CEDAR CITY
TROPIC
N
0
10
20
KM
ST. GEORGE
KANAB
114°
113°
112°
37°

Figure 9.

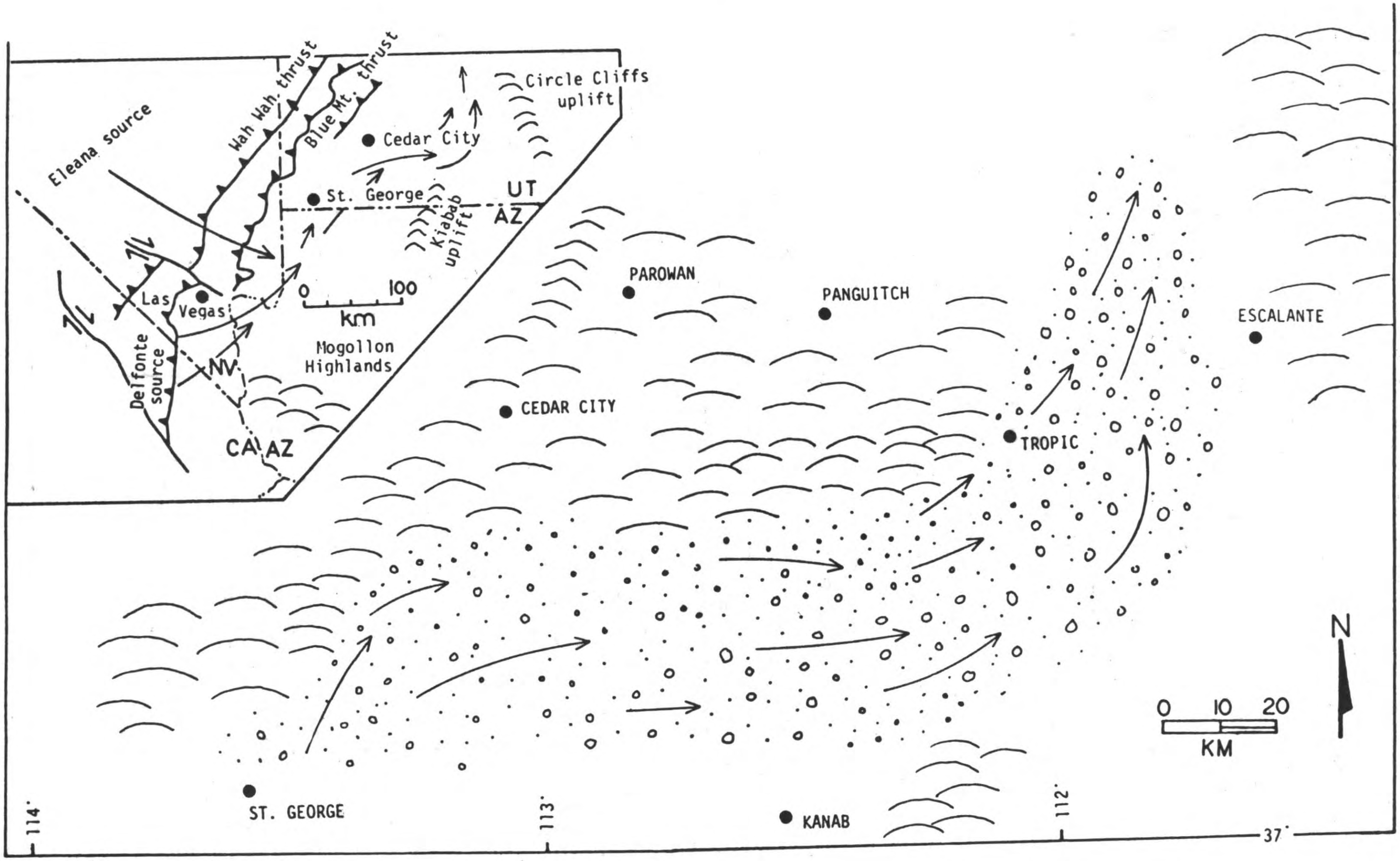

Eleana source
Wah Wah thrust
Blue Mt. thrust
Circle Cliffs uplift
Cedar City
St. George
UT
AZ
Kiabab uplift
Las Vegas
0 100
km
Mogollon Highlands
Delfonte source
NV
CA
AZ
PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
ST. GEORGE
KANAB
114°
113°
112°
37°
N
0 10 20
KM

Figure 10.

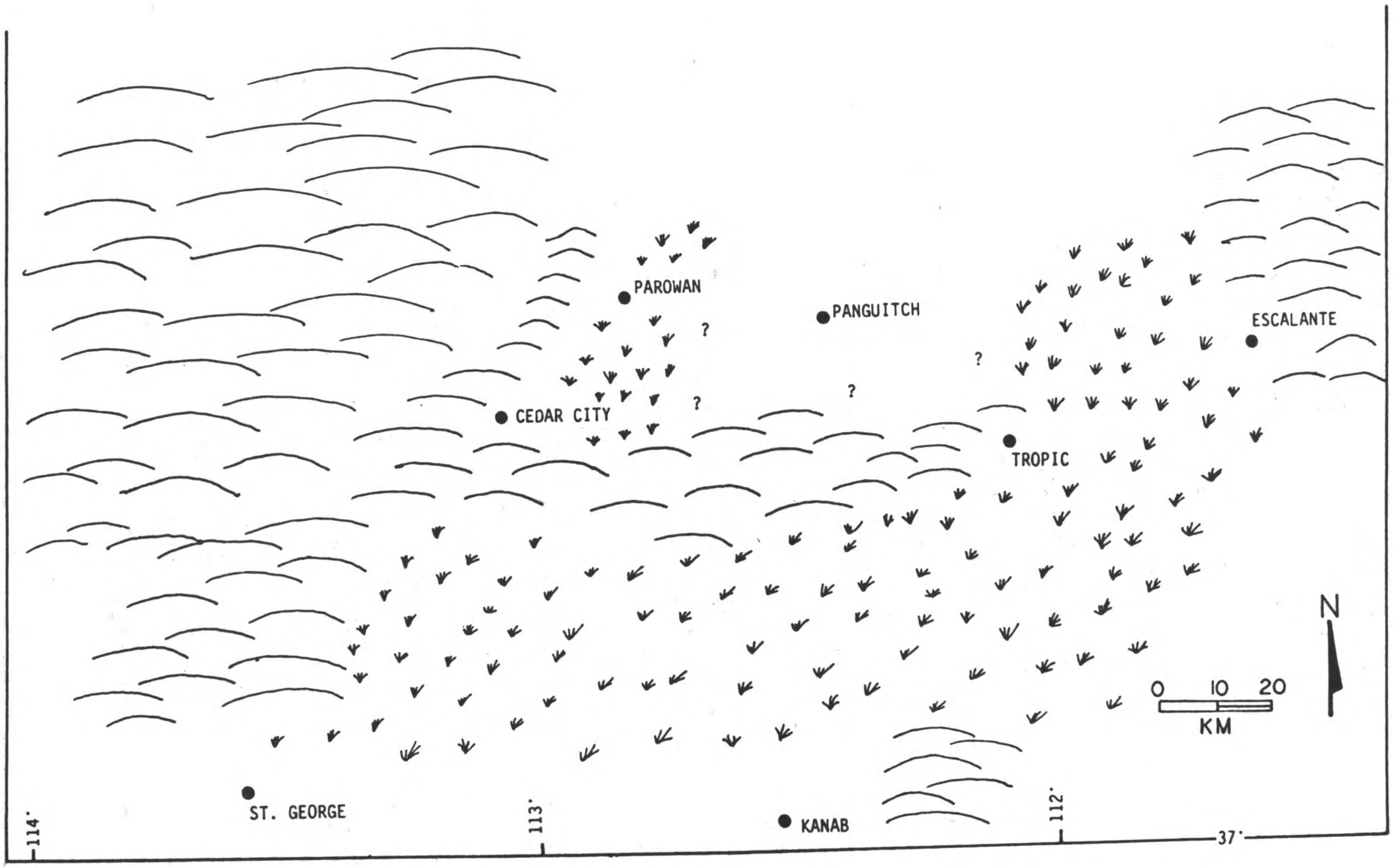

PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
N
0 10 20
KM
ST. GEORGE
KANAB
114°
113°
112°
37°

Figure 11.

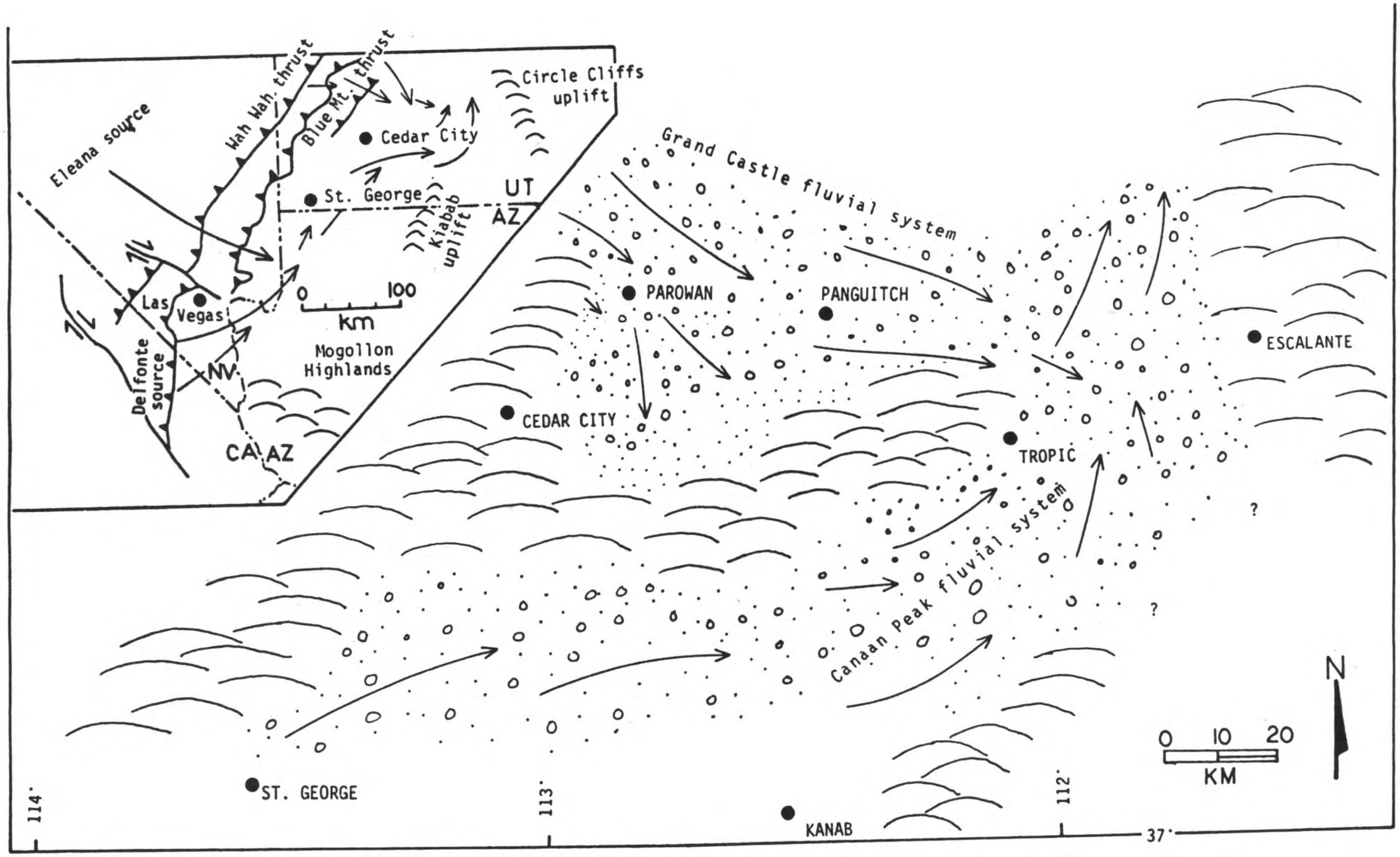

Eleana source
Wah Wah thrust
Blue Mt. thrust
Circle Cliffs uplift
Cedar City
St. George
UT
AZ
Kiabab uplift
Las Vegas
Delfonte source
NV
CA
AZ
0 100
km
Mogollon Highlands
Grand Castle fluvial system
PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
Canaan Peak fluvial system
ST. GEORGE
KANAB
114°
113°
112°
37°
N
0 10 20
KM
?
?

Figure 12.

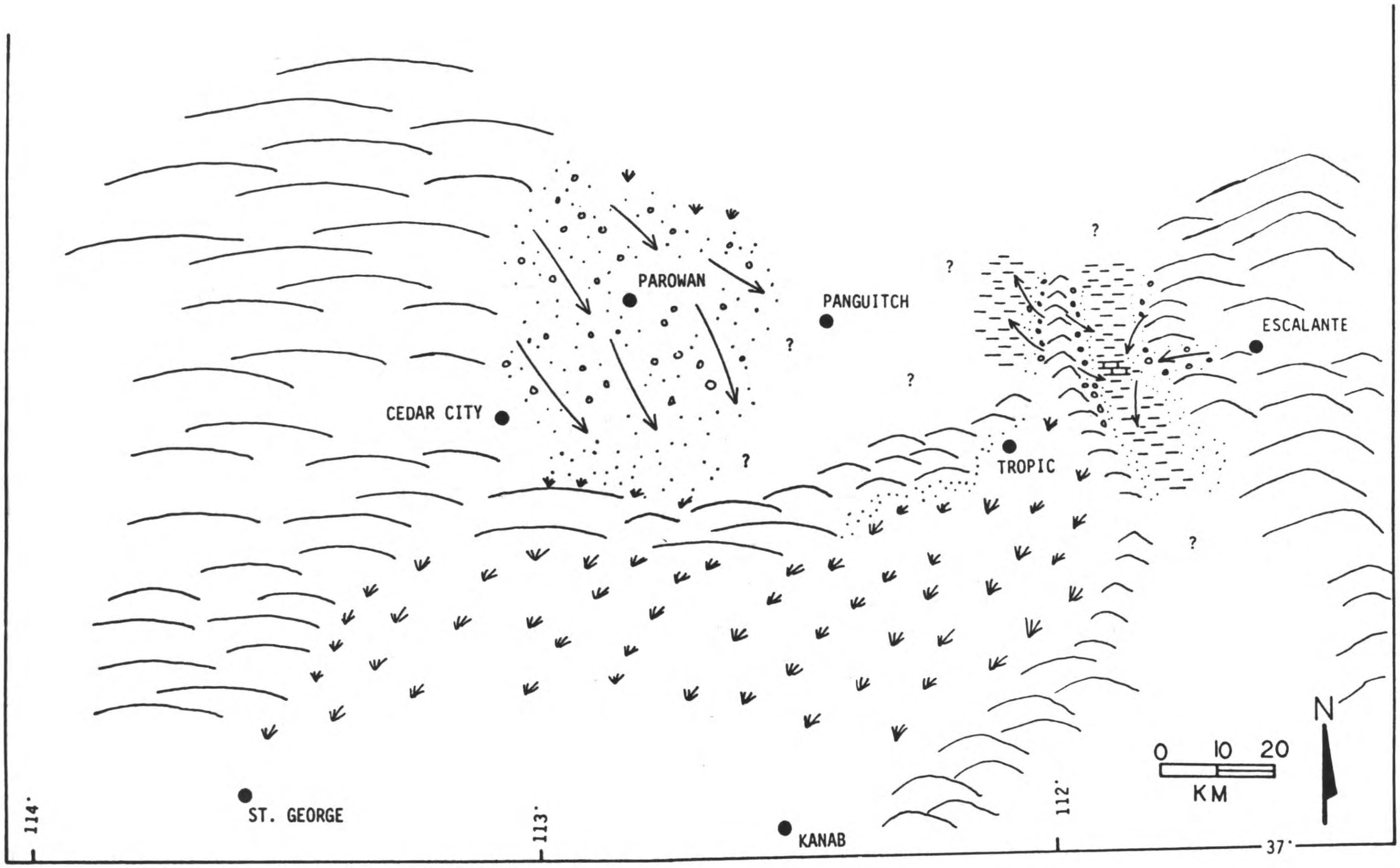

PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
ST. GEORGE
KANAB
114°
113°
112°
37°
0 10 20
KM
N
?

Figure 13.

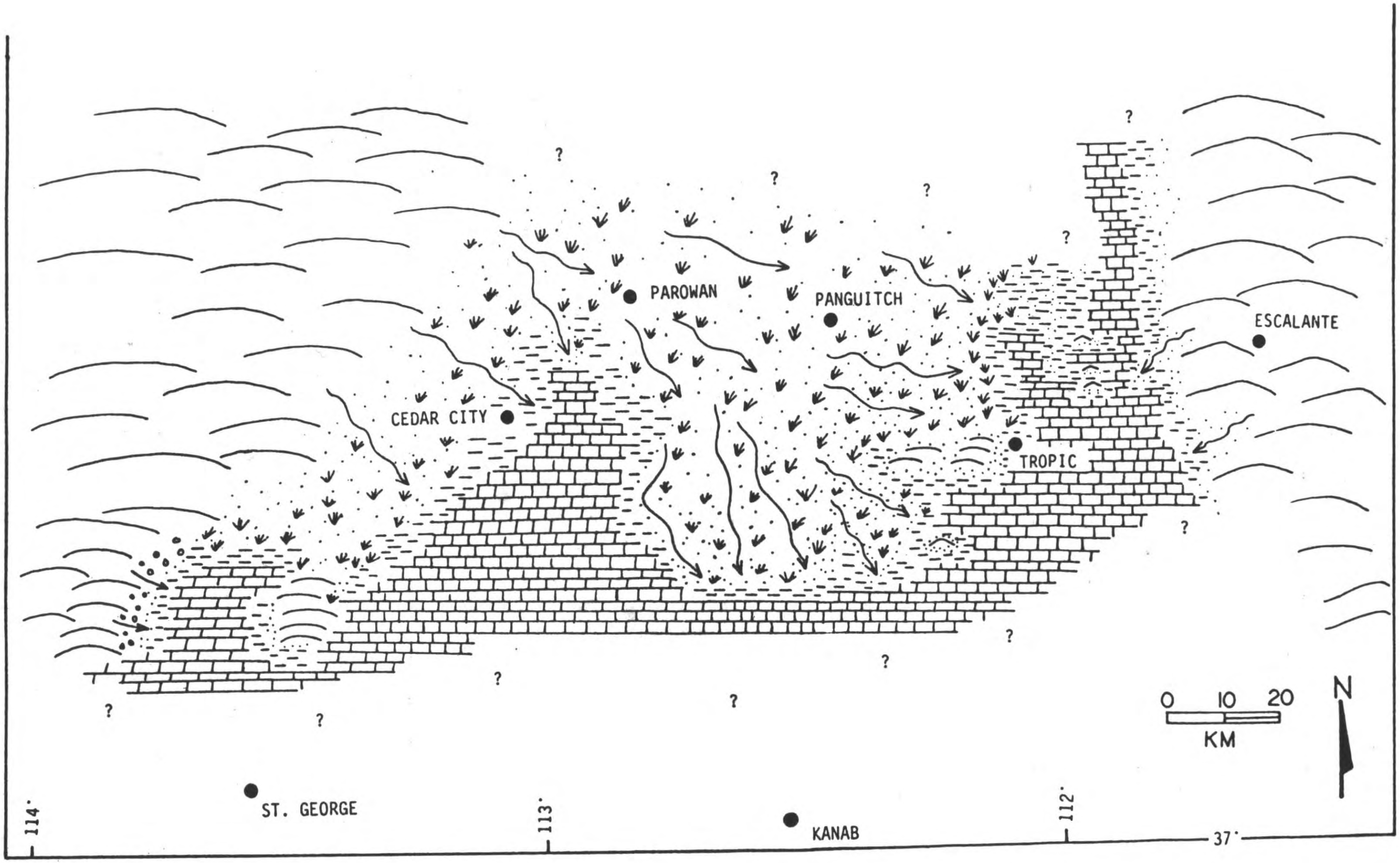

PAROWAN
PANGUITCH
ESCALANTE
CEDAR CITY
TROPIC
ST. GEORGE
KANAB
114°
113°
112°
37°
0 10 20
KM
N

Figure 14.